MODULE 2
BASIC BUILDING BLOCKS OF THUNDERSTORMS

OBJECTIVES

At the completion of this module, the student will be able to:

1) Describe how moisture, instability and lift contribute to thunderstorm development
2) Identify source regions for moisture
3) Differentiate between the various types of stability
4) Note ways in which the stability of an air mass can change
5) Identify some of the many causes of upward motion in the atmosphere
6) Understand the role of the capping lid in the development of thunderstorms

We all have our childhood memories of playing with blocks. Instinctively, we piled those blocks on top of one another in an effort to construct what was, in our minds, nothing short of a masterpiece. The concept of building blocks is fundamental to our study of the atmosphere as well. The so-called basic building blocks for thunderstorm formation are **moisture**, **instability**, and **lift**. Let's examine each to determine their role in the development process.

MOISTURE

In atmospheric studies, water, in its various forms, is a deep subject (no pun intended). It deserves great attention because it plays such an important role in weather generation. This is certainly the case in the development of thunderstorms. Adequate moisture is necessary for the formation of clouds and subsequent thunderstorms, and thus is considered one of the basic thunderstorm building blocks. In the interest of time and space, the concept of moisture and the role it plays in the thunderstorm development process can best be explained in the context of cloud formation and growth.

Source Regions

Moisture that finds its way into the atmosphere has its source in some body of water; be it a lake, a river, or some larger body such as the ocean. The primary source region for low-level moisture that leads to cloud development in and around North Texas is the Gulf of Mexico. Higher level moisture that moves into the area, usually from the west or southwest, oftentimes is carried aloft from the Pacific Ocean and transported thousands of miles.

Persistent south wind is a result of low pressure over the Texas Panhandle and western Oklahoma in juxtaposition with higher pressure over the eastern part of the country. This south wind is a good transporter of low-level moisture from the Gulf of Mexico. When the wind blows from this direction for extended periods of time (i.e. several days), more and more moisture is drawn northward into the area, leading to the "muggies" (warm and damp conditions) that are so typical of the warm season across the area.

Other source regions for low-level moisture, albeit minor, are area lakes, rivers, and streams. While on a large scale their role as a provider of atmospheric moisture is thought to be insignificant, on a local scale, this is probably not the case. Oftentimes, cloudiness, and sometimes thunderstorms, are observed to develop in the vicinity of lakes and major river systems.

Low-level moisture can also be transported vertically to higher levels in the atmosphere. This process, if allowed to operate for an extended period of time, can lead to the development of extensive layers of moist air. When these layers exist in the presence of other contributing factors, the chances for significant rainfall over a given area are dramatically enhanced.

Evaporation

At this point, we know where to look for water, but how does it get into the air? Quite simply, through the process of **evaporation** whereby liquid water is transformed into water vapor.

To illustrate how evaporation operates, try to envision molecules moving within a body of water at a speed proportional to the water's temperature (i.e. the higher the temperature, the faster the molecules move). Eventually, the temperature will reach a level at which the molecular speed will be sufficient to allow some of the molecules to escape the surface of the water. This is evaporation in action. As the molecules escape, they carry heat with them. This heat is carried with the vapor in hidden form and is referred to as **latent heat** (see Figure 2-1). Latent heat is an important concept that we will visit again later in this module.

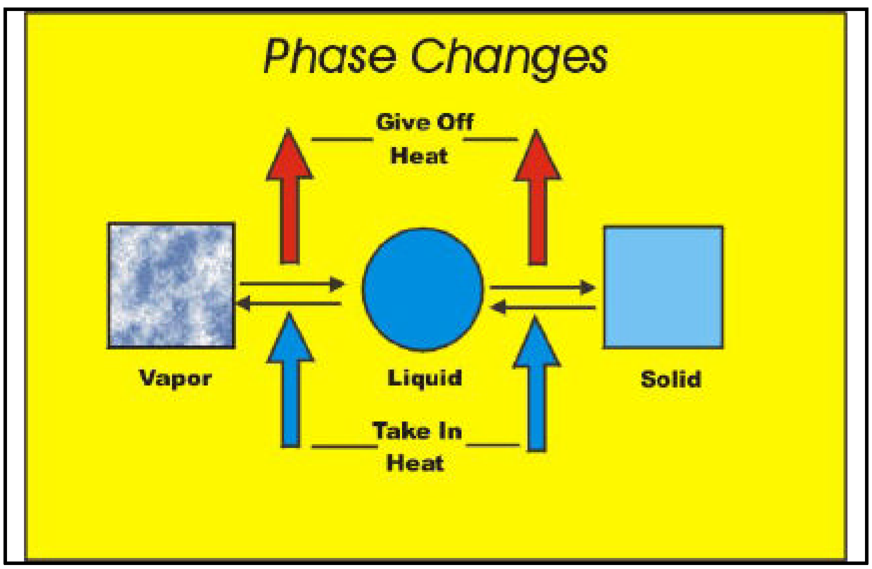

Figure 2-1: Phase changes of water and the associated cycle of heat storage and release.

As
evaporation continues and heat is extracted, the water cools. A vivid example of this is what happens when you step out of a swimming pool on a hot day. Water evaporates from your skin and you feel the chill.

Evaporation operates most efficiently when certain atmospheric conditions are prevalent. Warm temperatures increase the rate of evaporation. Also, dry air favors increased evaporation. Finally, windiness promotes evaporation as it carries water vapor away from a location, thus maintaining a strong moisture difference (gradient) between the water surface and the drier air above.

Moisture Measurement

At this point, we have the transformation of liquid water into water vapor occurring. As you might expect, the moisture content of the air is increasing. How do we measure it? There are a number of ways, but meteorologists most often measure the amount of moisture in the air in terms of its **dew point temperature**. This is the temperature to which air, at constant pressure, must be cooled in order to bring about saturation (see Figure 2-2). It is NOT affected by changes in the actual air temperature, unlike relative humidity, and thus is an accurate measure of the amount of moisture in the air (since fluctuations in air pressure at the surface are usually minor).

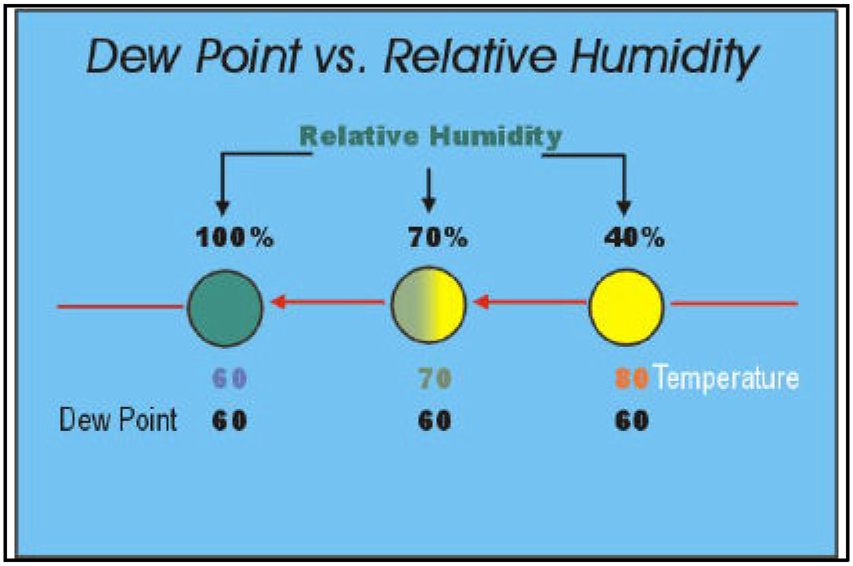

Figure 2-2: The relationship of dew point temperature to relative humidity.

High dew point temperatures, i.e. 60 degrees Fahrenheit or higher across North Texas during the spring storm season, indicate high moisture content of the air. Severe weather forecasters pay close attention to the movement of areas of high moisture. A surface pattern conducive to the horizontal transport, or **advection**, of high dew points into North Texas is shown in Figure 2-3.

Saturation

For the moment, we are going to assume that air over a given area is sufficiently moist in the lower levels and that conditions are favorable for parcels (small volumes of air) to rise, a process known as **convection**. As a parcel moves upward, it experiences less pressure from the ambient air, and as such, cools. Eventually, the temperature of the parcel may cool to its dew point temperature (since the air pressure is changing, the dew point actually decreases as well, but much less than the temperature decrease). When a parcel of air cools to its dew point, the air is said to be **saturated**, i.e. it cannot hold anymore water vapor. Important things happen when a parcel becomes saturated; things that we will discuss shortly.

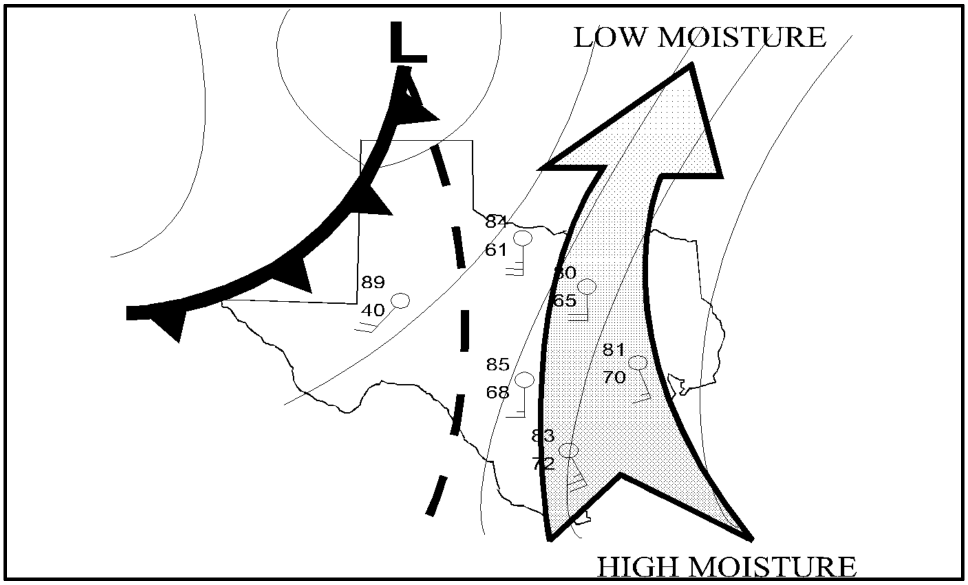

Figure 2-3: Surface pressure pattern conducive to the spread of high dew point (moisture) northward from the Gulf of Mexico.

For the time being, recognize that lowering a parcel's temperature is but one way to bring about saturation. Another way is to keep the temperature constant and add more water vapor to the parcel. In other words, evaporate more water into the parcel. Both these mechanisms often are in operation simultaneously.

Condensation

When a parcel reaches saturation and no longer has the capacity to hold additional water vapor, it attempts to reduce its water vapor load. What mechanism is available for the parcel to accomplish this? It transforms some of the vapor it contains into liquid water; a process known as **condensation**.

How successful the parcel is in condensing the vapor is primarily dependent upon the availability of particles (e.g. dust, salt, combustion byproducts) in the air on which water droplets can form. These particles are known as **condensation nuclei** and are quite small. Some of the particles attract water and promote condensation before the air actually reaches saturation. Others inhibit water droplet formation, even when the air is supersaturated.

Condensation results in cloud formation; the first step toward thunderstorm development. Remember the heat that was carried latent as a result of the evaporation process? Well, it's back. During condensation, that heat is released to the surrounding air (see again Figure 2-1). Thus, condensation can be thought of as a heating process and the release of this heat is vitally important in cloud growth and thunderstorm maintenance.

We'll postpone a discussion of cloud growth, and precipitation formation and fallout for the time being. We next examine the role of stability or more accurately, as in the case of thunderstorm development, instability.

INSTABILITY

The second basic building block for thunderstorm formation is **instability**. In a general sense, instability describes a condition in which there is the tendency for an air parcel to move away from its original position once it is disturbed. This is in contrast to **stability** in which the parcel, forced to move, returns to its original position. We'll discuss these concepts in more detail shortly. Before we can truly understand these concepts, we must delve deeper into moisture and temperature considerations.

Parcel Processes

Imagine our air parcel being completely isolated from its surroundings. Much like a balloon whose thin membrane prevents air inside from mixing with air outside, our parcel will not be allowed to mix with ambient (environmental) air. Any temperature changes to the parcel are a result of energy adjustments within the parcel. This condition, requiring there be no heat exchange between the parcel and its surroundings, defines an **adiabatic** process and is very important in cloud growth and thunderstorm development.

Despite the fact that our parcel will not be allowed to mix with the surrounding air, we must nonetheless consider its temperature in relation to that of the ambient air. Temperature change with height in the atmosphere is known as **lapse rate**. The environmental lapse rate is measured by sending instrumented balloons (radiosondes) upward through the atmosphere. Sensors measure the temperature at various levels and send the readings back electronically to a ground receiving station. The readings are then plotted to give a vertical profile of how temperature is changing with height. The normal lapse rate is a decrease of about 6.5 degrees C per 1000 meters (3.5 degrees F per 1000 feet).

Again considering our parcel, as it rises or sinks in the atmosphere, it too changes its temperature at a prescribed rate depending upon its moisture content. If the parcel is not saturated, its temperature changes at a rate of 10 degrees C per 1000 meters (5.5 degrees F per 1000 feet). This is the **dry adiabatic lapse rate**. As long as a parcel is not saturated, even if it is at 95 percent relative humidity, it will follow the dry adiabatic lapse rate. If the parcel can be cooled to its dew point, it becomes saturated. With saturation comes the release of latent heat. This addition of heat to the parcel slows the rate of temperature change slightly. The **moist adiabatic lapse rate** is 6 degrees C per 1000 meters (3.3 degrees F per 1000 feet).

Stability determinations in the atmosphere are made by comparing the environmental temperature to the temperature of the parcel at any height. Knowing these values and understanding that **warm air rises** and **cold air sinks** allows us to now consider the types of stability in the atmosphere.

Types of Stability

We can illustrate the concept of stability types by examining the movement of a ball on a curved surface (see Figure 2-4). Notice in the lower part of the figure that when a force is applied to the ball (in this case, a push), the ball moves away, but quickly returns to its original position. Of course, in the real world, the ball would oscillate about its original position until all of its energy was expended, but would nonetheless settle at its original position. This condition is known as **stable**, and, believe it or not, does play a role in thunderstorm development, as we will see.

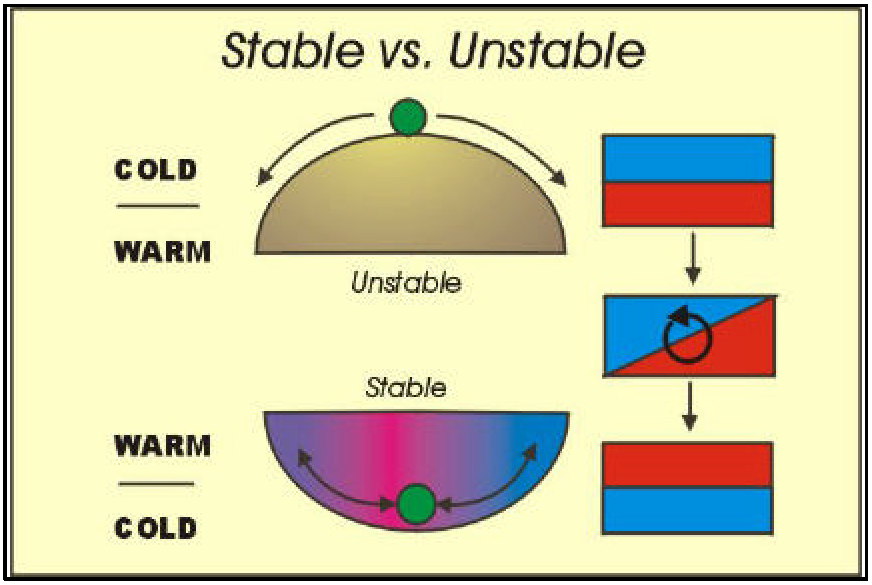

Figure 2-4: Illustration of the various types of stability.

If a push is given to the ball and the ball moves away and soon settles at a new location, we say that the condition is characterized by **neutral** stability. The ball will move only as long as there is a force moving it. Take the force away and the ball soon stops.

Instability is illustrated in the upper part of the figure. A force or push is applied to the ball and the ball continues to move further away from its original position. The ball will continue to move away without any additional application of force, as long as we maintain a condition of instability (i.e. we don't change the shape of the curved surface to that of a bowl or straight line).

We can now relate this illustration to the real world taking into account the warm and cold air laws we just mentioned. If a rising parcel suddenly finds itself colder than the environmental temperature at that height, it will begin to sink under its own weight as long as it remains relatively colder. This is a condition of stability. Upward vertical motion is inhibited.

If, on the other hand, the parcel moves into a region in which it is suddenly warmer than the surrounding environmental air, it will continue to rise as along as it remains warmer. This is a condition of instability. Severe thunderstorm forecasters examine vertical temperature profiles (soundings) to locate deep layers in which rising parcels will remain warmer than the ambient air and thus rise unabated.

Figure 2-5 illustrates the relationship between the environmental and process (parcel) lapse rates quite well. In (a), notice that a steep environmental lapse rate (one in which the air cools very rapidly with height) is unstable for both dry and moist parcels (the parcels will everywhere remain warmer that the environment). In (b), the drop off in environmental temperature is not very rapid and air parcels will everywhere remain colder than the environment and thus sink (stable to both dry and moist parcels). A third and important condition is shown in (c). A condition in which a dry parcel is stable (colder than the environment), but becomes unstable (warmer than the environment) once the parcel is saturated is known as **conditionally unstable**. This occurs quite often in environments where thunderstorms develop. Basically, all that is needed to release the instability and cause the thunderstorm to develop is some mechanism to mechanically force or lift the air to a height where it can cool to saturation, release latent heat, and become warmer than the environment. We'll examine some of these lifting mechanisms in the last section of this module.

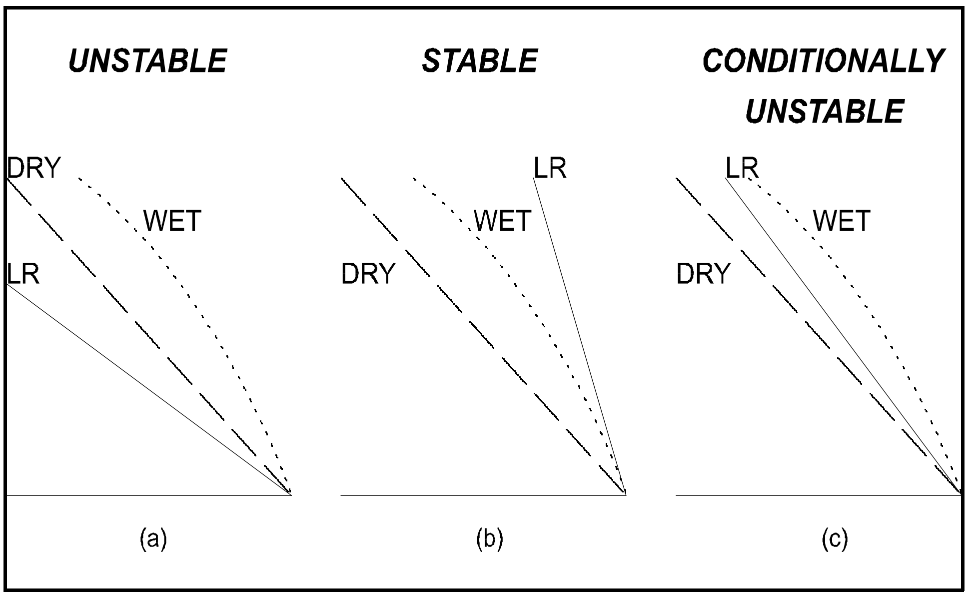

Figure 2-5: Relationship among the environmental lapse rate (solid), dry parcel lapse rate (dashed), and wet parcel lapse rate (dotted) for the various types of stability. Temperature increases to the right at the bottom of the graphs.

One other consideration should be made regarding rising and sinking air parcels. As air rises, it cools and expands. Why? As you'll remember, pressure in the atmosphere decreases exponentially with height. A rising parcel will experience less pressure as it moves up through the atmosphere (pressure is nothing more than the weight of the air in a column above a certain point; so, a rising parcel moves further away from the base of the air column). Air within the parcel is allowed to push outward with decreasing resistance. However, to do so, it must use some of its own energy which lowers its temperature. To see this process in action, try letting a little compressed air out of an automobile tire. How does the escaping air feel?

There is a flip side to this law. Sinking air comes under increasing pressure (moving closer to the base of the air column). Instead of being allowed to push outward, the increased outside pressure overpowers the air inside the parcel and actually compresses it. When you pump up a tire, what does the barrel of the pump feel like after about 20 or so compressions? Pretty warm, isn't it?

Sinking air warms and compresses.

Changing the Stability

We are now in a position to consider ways that the stability of a layer of air can be changed. The optimum condition for severe thunderstorm development, of course, is to change toward a condition of instability; i.e., cold air that wants to sink overlying warm air that wants to rise. Figure 2-6 shows this optimum condition.

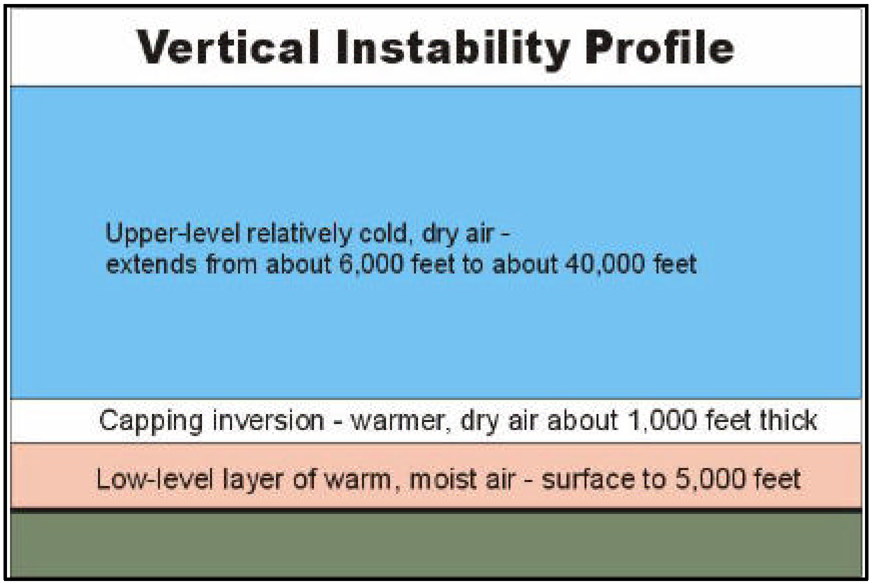

Figure 2-6: A typical vertical atmospheric profile that characterizes instability.

We must examine ways to warm the bottom of a layer, cool the upper part of the layer, or do both simultaneously. Radiation from the sun heats the ground up much faster than the air above it. But the heat that the ground stores up gets re-radiated to heat the air just above it. This is one way to warm the lower layers of the atmosphere.

Warm air can be brought in horizontally with the wind from another location (off the hot and dry plateaus of northern Mexico during the warmer months, for example). This process, referred to as **advection,** operates quite well when intense low pressure systems develop over West Texas. Strong south or southwest wind develops under these conditions bringing warm air rapaidly into the area.

Again, as we discussed, the air can be warmed through compression. Air that flows downhill (e.g. eastward from the higher terrain over West Texas and New Mexico) comes under increasing pressure and warms. This is referred to as **downslope** flow.

Finally, the addition of moisture to the air mass can result in warming. As the moisture condenses into cloud droplets, significant amounts of latent heat are released, warming the air.

We can independently or simultaneously cool the upper portions of a layer. The most obvious and perhaps frequent method is through advection of air from an upstream cold region. However, lifting, if allowed to operate long enough, can cause significant cooling, especially in conjunction with drying of the upper portions of a layer. Moisture lifted through a dry layer will evaporate which, as you'll remember from the swimming pool analogy, is a cooling process.

So, any of these processes working by themselves or in tandem can enhance the temperature difference across the layer. Severe thunderstorm forecasters refer to this process as **destabilization**.

Let's now look at our final building block: lift.

LIFT

We have adequate moisture, especially in the lower levels of the atmosphere, and there is instability through a deep layer. The stage is set. What sets the whole thing in motion? This is where lifting mechanisms come into play and there are a number of them to consider.

Factors Affecting Upward Motion

Let's first look at lift in a general sense. The atmosphere has some very sound laws that it cannot/ will not violate. Tops on this list of laws is that the flow of air around the globe is continuous, i.e. there can be no "holes" created by the air. Air flowing to one location must be replaced by air flowing from another. This is true both for air flowing horizontally and for air rising or descending vertically.

Keeping this in mind, let's examine Figure 2-7 which shows an idealized condition that supports continued lift. In the lower levels, air from different sources is coming together, a process known as **convergence**. Air that converges near the surface has few options for continued movement. Since the ground acts as a solid boundary preventing downward flow into it, the air must rise along this zone of convergence.

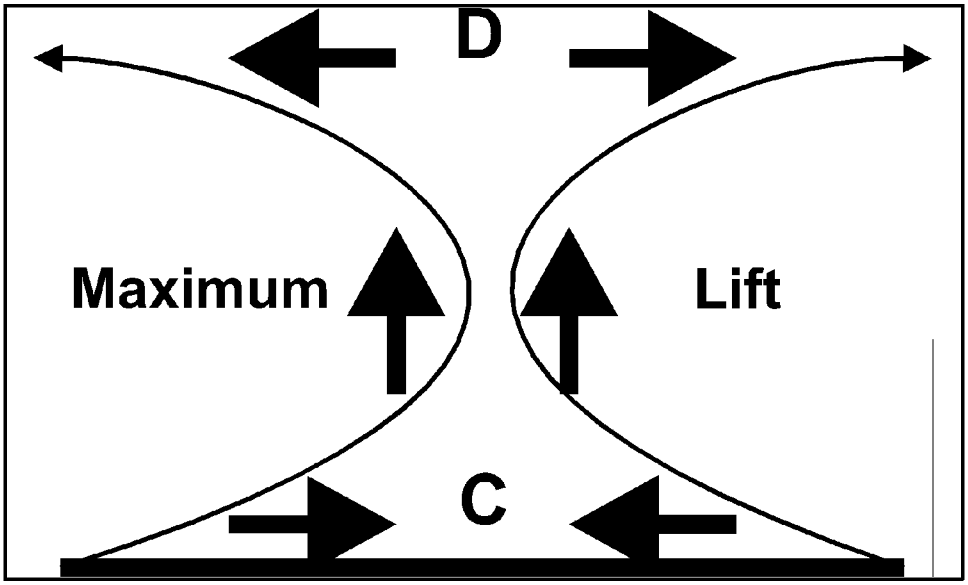

Figure 2-7. Schematic of how convergence in the lower levels (denoted by “C”) and divergence in the upper levels (denoted by “D”) promote continued lift.

Remembering that air that leaves a location must be replaced, it is easy to see how lifting of air leads to the establishment of a closed circuit of flow. Air comes together in the low levels, rises through some depth of the atmosphere and then flows apart (**divergence**). How well this pattern operates will determine the amount of upward motion that occurs.

If convergence at the surface dominates, air begins to pile up near the base of the air column and pressures rise. This impedes the lifting process. If, on the other hand, divergence aloft out-performs low-level convergence, air is evacuating the column causing pressure to decrease at the base of the column. This leads to stronger inflow (greater convergence) to compensate (remember, the atmosphere likes continuity).

Another factor to consider in the convergence/divergence process is the degree of atmospheric instability. Air that is forced to rise under unstable conditions will accelerate upward. If upper-level divergence can operate as effectively or more effectively than lower-level convergence, a well-developed vertical motion cycle is established.

The vertical motions caused by convergence/divergence patterns such as this usually take place over a large area, but aren’t strong enough to initiate thunderstorm development on their own. These large-scale vertical motions can help destabilize the atmosphere and enhance any storms that are able to develop by other means. For storm development, we generally look for sources of stronger low-level lift.

Sources of Storm-Scale Upward Motion

We will start our discussion of lift by describing the role that surface heating plays. We know from previous discussion that warm air rises and cold air sinks. Ground areas that heat up more than adjacent areas will be regions where parcels of warm air break away and rise through the relatively cooler surrounding air. Air must be brought in from the surrounding area to compensate for the loss of air due to lifting. For localized regions, heating-induced convection/convergence can operate quite well and initiate thunderstorm development.

The process of convection can be illustrated by observing a pot of water heating on a stove. Bubbles begin to form on the bottom of the pot and rise through the surrounding cooler water. Due to the laws of continuity, the surrounding relatively cooler water will sink in the pot creating a vertical circulation pattern.

Convection transports heat from the lower levels upward and directs some of the cooler air aloft downward. Rising parcels are generally invisible until condensation occurs. For a given area, condensation of parcels takes place at essentially the same level; thus, the common observation of seeing a cluster of clouds with bases at the same level. Below the base upward parcel speeds can be fairly strong, on the order of 15-20 mph (8-10 m/s). As clouds grow taller, the updrafts become more intense. Updraft speeds can increase to 40-50 mph (20-25 m/s) and higher.

Small-scale boundaries can produce more spatially-extended lift. The two most common of these lifting mechanisms are the **gust front (outflow boundary)** and **sea breeze front**.

The sea breeze front (Figure 2-8) operates on the principle just discussed. At coastal locations, differential heating of land and water causes the air over land to warm more rapidly and rise. Air rushes in from over the water to replace the rising air, creating a thermally-driven circulation where air in the lower-levels converges and lifts over land and sinks and diverges over the water. The leading edge of this circulation, the sea breeze front, can trigger thunderstorm development. The environment near sea breezes is generally not favorable for long-lived or severe thunderstorms. Such fronts can, however, interact with other boundaries and lead to more significant outbreaks of weather.

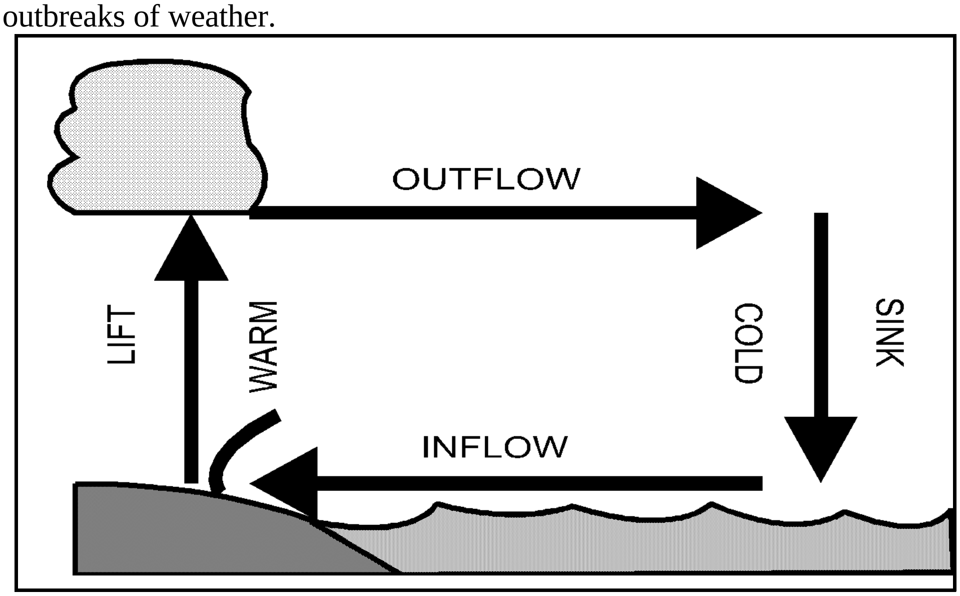

Figure 2-8: Schematic of the sea breeze circulation and front.

Gust fronts, often referred to as outflow boundaries, are produced by downward and outward flows of air from thunderstorms (Figure 2-9). The downward-moving air from deep within a thunderstorm can be quite cold. This cold, dense air accelerates toward the ground and spreads out horizontally along the surface. The interface between this outward-flowing air and the ambient air often flowing toward the thunderstorm forms a frontal-like boundary that can trigger new thunderstorm development. Again, the local environment in which this new lift occurs will determine the severity of any storms that form.

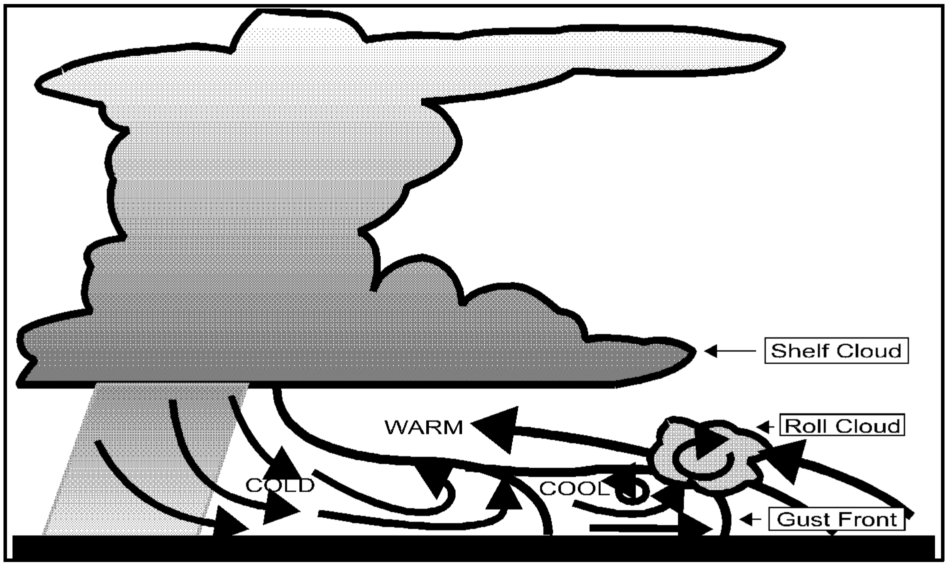

Figure 2-9: Cross-sectional view of a gust front (outflow boundary).

The analysis in Figure 2-10 depicts a larger-scale boundary known as a **dryline**. The dryline is notorious for its ability to trigger strong upward motions that lead to severe thunderstorm development. In fact, it probably is the most frequent initiator of severe thunderstorms that affect the Dallas-Fort Worth Metroplex.

The dryline is strange in that it “mixes” (via convective processes) to the east during the day and “advects” (with the wind) back to the west during the night. How rapidly the dryline can “mix” to the east is dependent on the amount of moisture it must “mix out”, and therefore the depth of the moisture layer. Basically, the dryline represents a line of demarcation between warm, moist air to its east and hot, dry air to the west. By virtue of the hot air being extremely light, it flows up and over the denser and relatively cooler marine air. Figure 2-11 shows a typical vertical profile of the dryline. To the west, the moisture layer is quite shallow and very little daytime heating of the lower levels is required to mix the drier air down to the surface. Thus, the dryline will almost appear to “leap” to the east. As the boundary interacts with deeper moisture farther east, more time is needed for the mixing process to accomplish its goal, and the dryline appears to slow down. Eventually, the moisture depth is too extensive to mix out before daytime heating ends in the evening and eastward progression of the dryline boundary ceases.

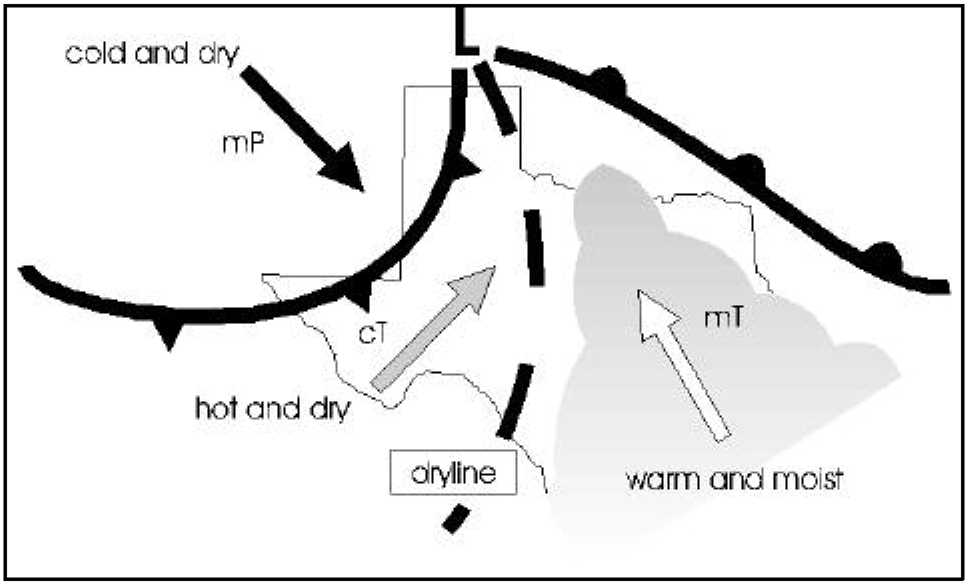

Figure 2-10: Surface analysis depicting the typical position of the dryline.

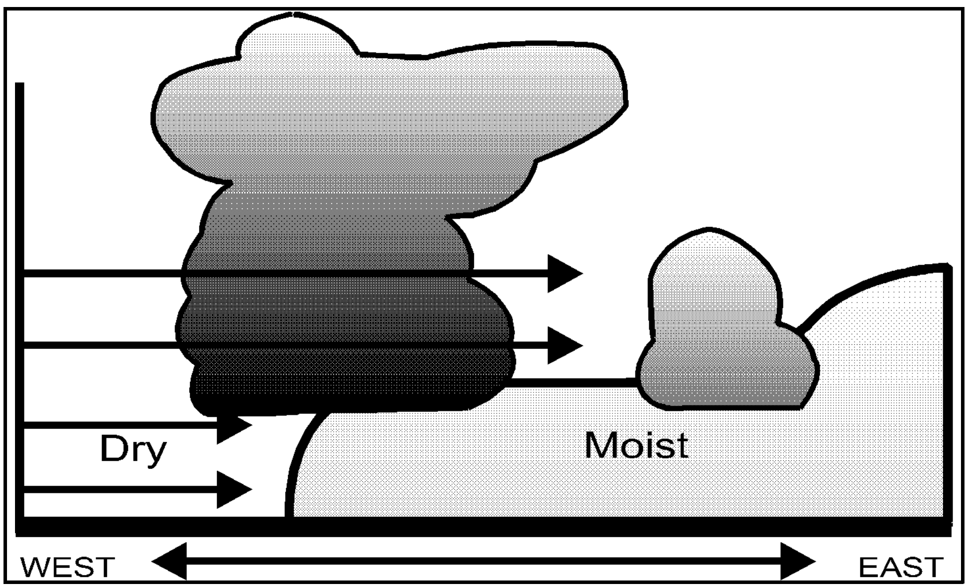

Figure 2-11: Vertical moisture and temperature profile in the vicinity of the dryline.

Overnight the dryline treks back to the west. However, the transport mechanism is different. The motion of the dryline westward is a function of the wind in the lower levels. Advection drives the process. East and southeast winds bring moisture inland from the Gulf of Mexico that literally "pushes" the dryline westward.

Convergence of wind and moisture is a common feature in the vicinity of the dryline. When this convergence occurs in the presence of other conditions that favor severe thunderstorm development (e.g. instability and strong winds aloft), it becomes a major triggering mechanism. During late Spring and early Summer, the dryline may become established for several days in a row bringing North Texas an almost daily threat of severe weather.

The boundaries of largest scale (often several hundred miles in length) that act as triggers for thunderstorm development are the true surface fronts. Fronts typically separate air masses on the basis of temperature contrast, as opposed to moisture with the dryline. Fronts are classified as warm, cold, or stationary on the basis of the movement of the cold air; if advancing southward, cold; if retreating northward, warm; if essentially no movement, stationary.

Like the dryline, fronts provide a mechanical means of lift. Figure 2-12 shows how cold air pushes underneath warmer air creating a steeply sloping boundary. Warm and moist air glides up this boundary far enough to produce clouds. If the atmosphere is unstable and lift can be sustained to a point at which the warm cloud parcels can rise on their own without further mechanical input, thunderstorm development becomes a distinct possibility.

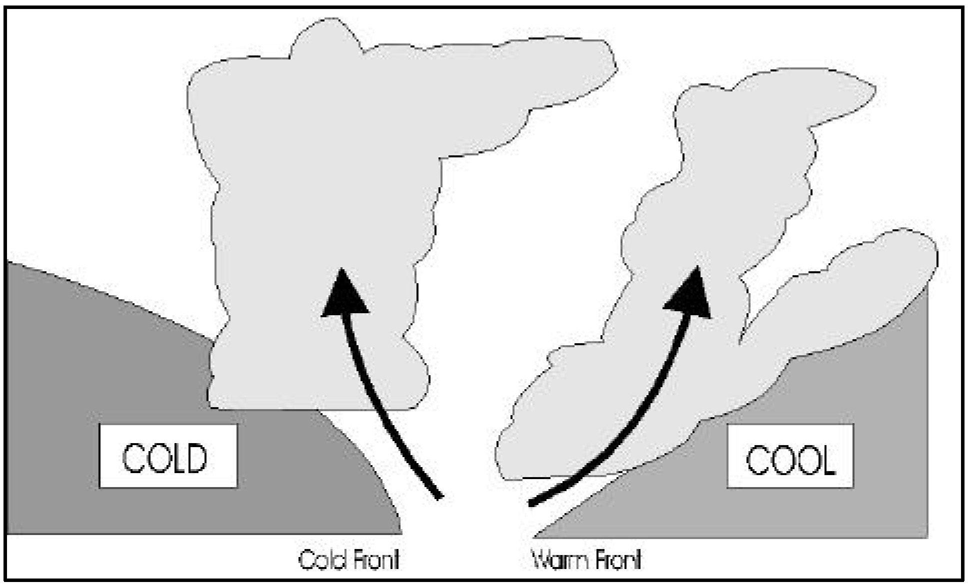

Figure 2-12: Cross-sectional view of a frontal boundary showing the undercutting of cold air and the overriding of warm air.

In the Dallas-Fort Worth area, a significant number of thunderstorms occur as a result of lifting along a front. However, the weather systems that drive fronts through our area also favor dryline development. Thus, an even greater percentage of thunderstorms, especially severe thunderstorms, develop as a result of dryline lift.

We need to also consider the effects of orographic features on the production of lift. In Figure 2-13, a hill or mountain is placed in the path of horizontal wind flow. Since air on the upwind side of the obstruction cannot flow through the side of the obstruction, it must flow around or over it.

Moist air that is forced to ascend will eventually form clouds. Again, instability plays a major role in determining the cloud's potential for growing into a thunderstorm. With sufficient instability present, persistent flow toward the orographic feature will maintain a pattern of cloud and thunderstorm development along and near the feature.

As the storms grow, upper level winds may carry them away from the source region for lift, which may result in weakening. Indeed, this cycle occurs quite often along the eastern foothills of the Rocky Mountains (refer back to Figure 1-1). Storms develop sometimes quite rapidly under conditions of low-level east or southeasterly wind and strong daytime heating. If mid and upper-level instability is present, some of the storms can become severe. However, upper-level southwest wind will oftentimes carry these storms away from the mountains where the low-level lift is concentrated and cause them to weaken before penetrating much into the western plains.

The downwind side of the orographic feature, referred to as the **leeside**, is an area where flow descends. In addition to the moisture lost in condensation on the windward side, descending air on the leeside warms and remaining moisture is lost to evaporation. Thus, little or no precipitation occurs on this side.

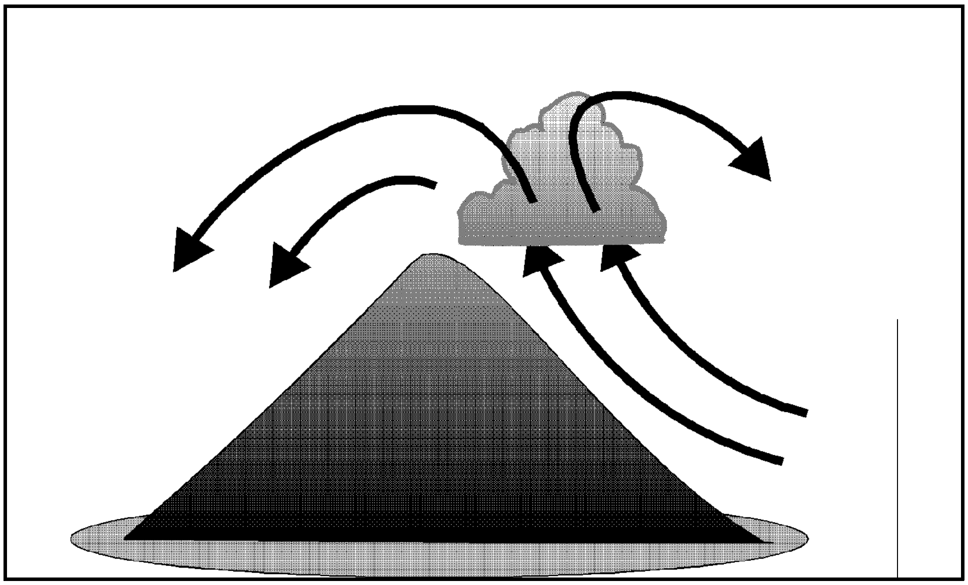

Figure 2-13: Lift produced by the presence of an orographic feature.

THE CAPPING LID

We have presented a rather detailed explanation of how thunderstorms form. However, despite having all of the necessary ingredients available, the development process can be thwarted by the presence of one inhibiting factor. Meteorologists refer to this limiting factor as the **capping lid**, or the **cap** for short. The capping lid is an elevated **inversion** (shown in Figure 2-14), a stable layer in which temperature increases rather than decreases with height, of sufficient strength (stability) to prevent the initiation or continuation of upward motion. Typically, the cap is found at an altitude of about 10,000 feet. When the atmosphere is “capped”, all of the ingredients that are in place to produce severe thunderstorms are “trapped” and unable to operate.

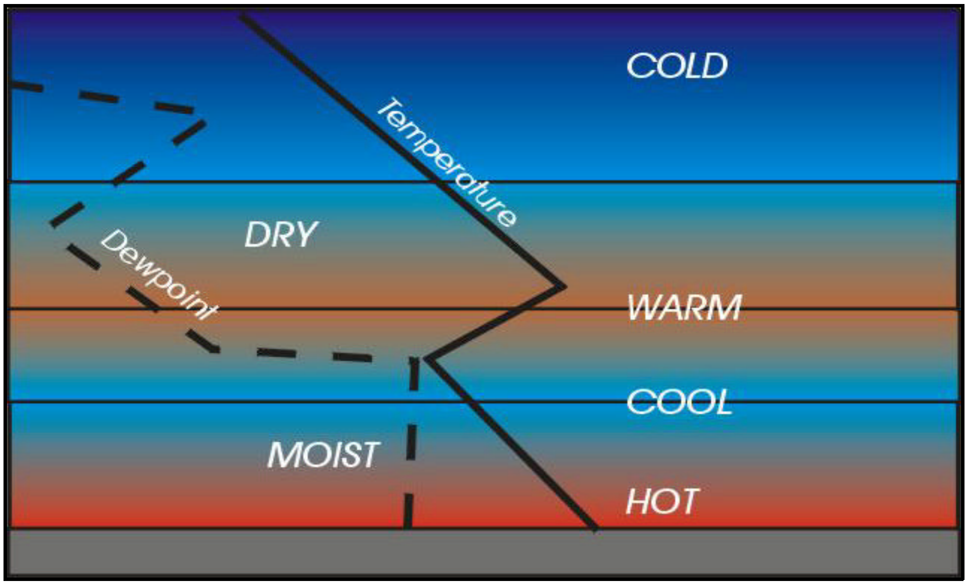

Figure 2-14: The capping lid. The warm layer serves to keep thunderstorm development from occurring until the air beneath it becomes very hot and/or moist.

What appears negative at first glance does, however, have positive effects for the production of severe thunderstorms. Obviously, to set things into motion, something must come along to remove this lid. A trigger or lift mechanism of sufficient strength (e.g. dryline) often does the trick. But until then, the atmosphere below the inversion increases in explosive potential, i.e. it gets warmer and more moist, and therefore more unstable. Once the lid is removed, development can be extremely rapid with particularly violent updrafts occurring.